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Isotopic constraints on the age and early differentation of the Earth

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Abstract

The Earth's age and early differentiation history are re-evaluated using updated isotopic constraints. From the most primitive terrestrial Pb isotopic compositions found at Isua Greenland, and the Pilbara of Western Australia, combined with precise geochronology of these localities, an age 4.49 ± 0.02 Ga is obtained. This is interpreted as the mean age of core formation as U/Pb is fractionated due to sequestering of Pb into the Earth's core. The long-lived Rb-Sr isotopic system provides constraints on the time interval for the accretion of the Earth as Rb underwent significant depletion by volatile loss during accretion of the Earth or its precursor planetesimals. A primitive *measured* $^{87}Sr/^{86}Sr$ initial ratio of 0.700502 ± 10 has been obtained for an early Archean (3.46 Ga) barite from the Pilbara Block of Western Australia. Using conservative models for the evolution of Rb/Sr in the early Archean mantle allows an estimate to be placed on the Earth's initial Sr ratio at ≈4.50 Ga, of 0.69940 ± 10. This is significantly higher than that measured for the Moon (0.69900 ± 2) or in the achondrite, Angra dos Reis (0.69894 ± 2) and for a Rb/Sr ratio of ≈1/2 of chondrites corresponds to a mean age for accretion of the Earth of 4.48 ± 0.04 Ga.

The now extinct ^{146}Sm-^{142}Nd ($T_{1/2}^{146}$=103 10^6yrs) combined with the long-lived ^{147}Sm-^{143}Nd isotopic systematics can also be used to provide limits on the time of early differentiation of the Earth. High precision analyses of the oldest (3.8-3.9 Ga) Archean gneisses from Greenland (Amitsoq and Akilia gneisses), and Canada (Acasta gneiss) do not show measurable (> ± 10ppm) variations of ^{142}Nd, in contrast to the 33 ppm ^{142}Nd excess reported for an Archean sample. The general lack of ^{142}Nd variations, combined with the presence of highly positive ε_{143} values (+4.0) at 3.9 Ga, indicates that the record of large-scale Sm/Nd fractionation events was not preserved in the early-Earth from 4.56 Ga to ≈4.3 Ga. This is consistent with large-scale planetary re-homogenisation during ongoing accretion of the Earth. The lack of isotopic anomalies in short-lived decay systems, together with the Pb and Sr isotopic constraints is thus consistent with core formation and accretion of the Earth occurring over an ≈100 Ma interval following the formation of meteorites at 4.56 Ga.

Introduction

A fundamental question concerning the origin of the Earth is its age. Did the Earth accrete shortly after the formation of primitive meteorites at 4.56 Ga or was there a substantial interval before final accretion and early differentiation of the Earth? Conclusive answers to this question remain elusive, but a number of lines of evidence that are considered here now appear to indicate that the accretion of the Earth was not completed until at least 100 Ma after the formation of chondrites. The difficulty in establishing a precise age for the Earth arises from the lack of a geological record of the first ≈500 Ma of Earth history. The only known remnants from this period are 4.27 Ga detrital zircons found in an Archean sediment from the Jack Hills region of Western Australia (Compston & Pidgeon 1986). This provides a firm minimum age for the Earth, but the question of what happened between 4.27 Ga and 4.56 Ga remains. Insights into the Earth's earliest history can however be obtained from both the long-lived Rb-Sr, Sm-Nd and U-Pb systems as well as the now extinct ^{146}Sm-^{142}Nd decay series.

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The formation of the Earth by accretion of planetismals was undoubtedly a complex process involving large-scale melting, internal differentiation of a metallic core and formation of a crust. These processes were characterised by distinctive patterns of geochemical fractionation, and it is this time, or mean age of parent/daughter fractionation which is registered by isotope systems. In the case of the U-Pb system, Pb has as a strongly chalcophile character and as a result was probably strongly partitioned into the Earth's core. Thus the substantially higher (≈100x) U/Pb ratio for the Earth's mantle relative to carbonaceous chondrites is attributed to Pb being sequestered into the Earth's core along with sulphur. For this reason, the U-Pb system provides constraints on the average time of formation of the Earth's core. For the Rb-Sr system, the major cause of fractionation is the volatile nature of Rb compared to Sr. Thus the ≈10x depletion of Rb in the Earth compared to chondrites is attributed to volatile loss processes during accretion of the Earth. In contrast to these systems, Sm and Nd are refractory elements, and are strongly fractionated only during magmatic differentiation processes. The question of the 'age of the Earth' is therefore be considered from a number of different aspects; the time-scales for accretion, core formation and early mantle differentiation.

Pb isotopic constraints on core formation

The application of long-lived radiometric chronometers to the question of the Earth's age was pioneered by Rutherford (1929). Based on the present-day abundances of the long-lived isotopes of uranium ^{235}U and ^{238}U, he deduced an approximate estimate for the timing of stellar nucleosynthesis and hence a maximum age of the Earth's of ≈4 billion years. The first precise estimate of the Earth's age was made by Patterson (1956) who showed that the Pb isotopic evolution of the Earth and meteorites approximately corresponds to a single-stage evolution from about 4.55 Ga to the present. This conclusion was based on the close proximity of the average modern terrestrial Pb to a 4.55 Ga isochron defined by meteorites. The value of modern terrestrial Pb used by Patterson was derived from the average Pb isotopic composition of marine sediments in the Pacific ocean. Since this pioneering work, it has now become apparent that modern rocks have a much more diverse range of Pb isotopic compositions indicative of complex, multi-stage evolutionary histories. For example, Allegre *et al.* (1995) has shown that modern mid-ocean ridge basalts (MORB's) have a wide range of generally younger single-stage Pb-Pb ages of from 4.3 Ga to 4.6 Ga.

In order to critically address the question of whether the Earth is significantly younger than its meteorite parent bodies, it is useful to consider the initial Pb isotopic composition of the oldest terrestrial rocks. Primitive terrestrial Pb compositions provide the least equivocal record of U-Th-Pb fractionation following the Earth's formation and hence the most reliable isotopic constraints on the Earth's age and early differentiation history. For a single-stage model of Pb evolution, the relationship between the parent ^{238}U and ^{235}U and daughter ^{206}Pb and ^{207}Pb isotopes is given by;

$$\frac{^{207}Pb/^{204}Pb(t) - {}^{207}Pb/^{204}Pb(I)}{^{206}Pb/^{204}Pb(t) - {}^{206}Pb/^{204}Pb(I)} = \frac{^{235}U(e^{\lambda' T_{core}} - e^{\lambda' t})}{^{238}U(e^{\lambda T_{core}} - e^{\lambda t})} \quad (1)$$

where $^{207}Pb/^{204}Pb(I)$ and $^{206}Pb/^{204}Pb(I)$ are the initial Pb compositions at the time of core formation T_{core}. $^{207}Pb/^{204}Pb(t)$ and $^{206}Pb/^{204}Pb(t)$ are the initial Pb compositions in early Archean rocks of age t (Table 1). The decay constants for ^{238}U and ^{235}U are $\lambda = 0.155125$ Æ$^{-1}$ and $\lambda' = 0.98485$ Æ$^{-1}$ respectively, with $^{235}U/^{238}U = 1/137.88$. This equation enables the age of the Earth's core to be calculated using estimates of the Earth's initial Pb composition derived from meteorites, together with the most primitive Pb compositions preserved in ancient terrestrial rocks. Listed in Table 1 are samples with primitive Pb isotopic compositions; ie $^{206}Pb/^{204}Pb < 12$. An equally important constraint in the application of this equation is uncertainties in age of the early Archean initial Pb's (*e.g.* Gancarz & Wasserburg 1977). SHRIMP U-Pb dating of zircons has recently been undertaken at both Isua (Nutman *et al.* 1995) and in the Pilbara, for the Duffer dacite (McNaughton *et al.* 1993), the host of the Big Stubby galena. This improved chronology is particularly helpful in the interpretation of initial Pb's from the Isua supracrustal belt (Appel *et al.* 1978, Richards & Appel 1987). Nutman *et al.* (1995) have shown that this complex includes at least two felsic volcanic complexes with ages of 3.708 ± 0.003 Ga and 3.807 ± 0.002 Ga, with the galenas listed in Table 1 being associated with the older volcanics of the western section.

Following the approach of Gancarz & Wasserburg (1977), the initial Pb's listed Table 1, are presented, using plots of $^{207}Pb/^{206}Pb$ versus $^{204}Pb/^{206}Pb$ (Fig 1). This diagram emphasises the evolution of Pb in the early Archean. It is apparent from Figure 1b that plausible ranges for $^{238}U/^{204}Pb$ (μ) of from 8.5 to 9.5, are required to be consistent with the evolution of Isua, Big Stubby as well as younger conformable Pb's (Stacey & Kramers 1974). With this range of μ the Pb-Pb age of the Earth's core, T_{core} is estimated to be from 4.52 Ga to 4.46 Ga (Table 1). This age is clearly younger than that of chondritic meteorites (4.563 Ga), and hence the value of initial Pb given by Canyon Diablo (Tatsumoto *et al.* 1973) must be adjusted to take into account the Pb evolution from the time of meteorite formation, at 4.563 Ga to ≈4.50 Ga. The average μ for carbonaceous chondrites is low (< 0.2) and based on the correlation of Pb/U versus K/U, a $\mu_1 \approx 1$ is inferred (Allegre *et al.* 1995) for the proto-Earth prior to core formation. Thus the corrected initial Pb for the Earth at ≈4.50 Ga is given by;

$$^{206}Pb/^{204}Pb(I) = {}^{206}Pb/^{204}Pb(CD) + \mu_1(e^{\lambda 4.56} - e^{\lambda T_{core}}) \quad (2)$$

Table 1

Most primitive terrestrial Pb isotopic compositions and the age of the Earth's core

Locality	Age (Ma)	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	208Pb/^{204}Pb	T_{core}(Ga)	μ_2
ISUA						
261619[a]	3807 ± 2	11.176	13.047	31.130	4.492	9.11
263654[a]	3807 ± 2	11.311	13.147	31.267	4.419	11.0
2966[b]	3808 ± 2	11.146	13.025	31.148	4.510	8.72
Amitsoq[c]	3590 ± 50	11.468	13.203	31.365	4.462	8.46
PILBARA						
Big Stubby[d,e]	3465 ± 3	11.912	13.707	31.838	4.495	8.71
Canyon Diablo	4563[f]	9.307	10.294	29.476		
INITIAL Earth Pb						
	4490[g]	9.330	10.339	29.495	$\mu_1 = 1.0$	

[a] Richards & Appel (1987); [b] Appel *et al.* (1978); [c] Gancarz & Wasserburg (1977); [d] Richards (1986); [e] Pidgeon (1978); [f] Tatsumoto *et al.* (1973); [g] this study.

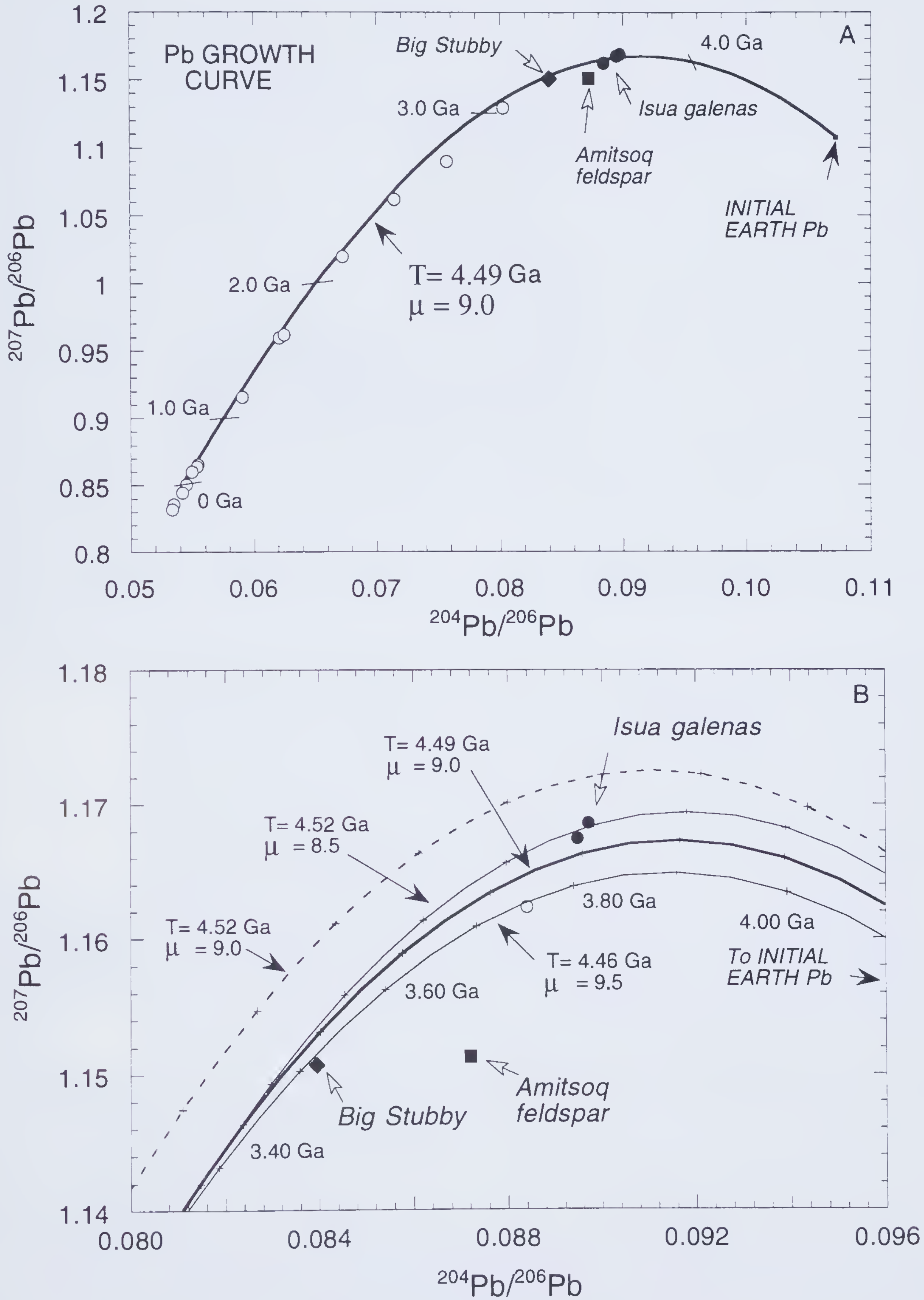

Figure 1A. Pb growth curve plotted using $^{206}Pb/^{207}Pb$ versus $^{204}Pb/^{206}Pb$. Primitive terrestrial Pb's are shown with solid symbols and open circles are galena Pb's tabulated by Stacey & Kramers (1974). Earth initial Pb from Canyon Diablo (Tatsumoto *et al.* 1973). **B.** Plot of the most primitive terrestrial Pb ($^{206}Pb/^{204}Pb < 12$) for galena from Isua, West Greenland (Appell *et al.* 1978; Richards & Appel 1987) and Big Stubby, Pilbara (Pidgeon 1978; Richards 1986). Amitsoq feldspar from Gancarz & Wasserburg (1977). Open circle shows disturbed younger sample from Isua (Richards & Appel 1987). The curves show the Pb evolution for μ = 8.5, 9.0 and 9.5. The best fit to both the Isua and Big Stubby Pb's is given for an age of the Earth's core of T_{core} = 4.49 Ga and u_2 = 9.0. Initial Earth Pb is evolved Canyon Diablo using μ_1 - 1 from 4.56 Ga to 4.49 Ga.

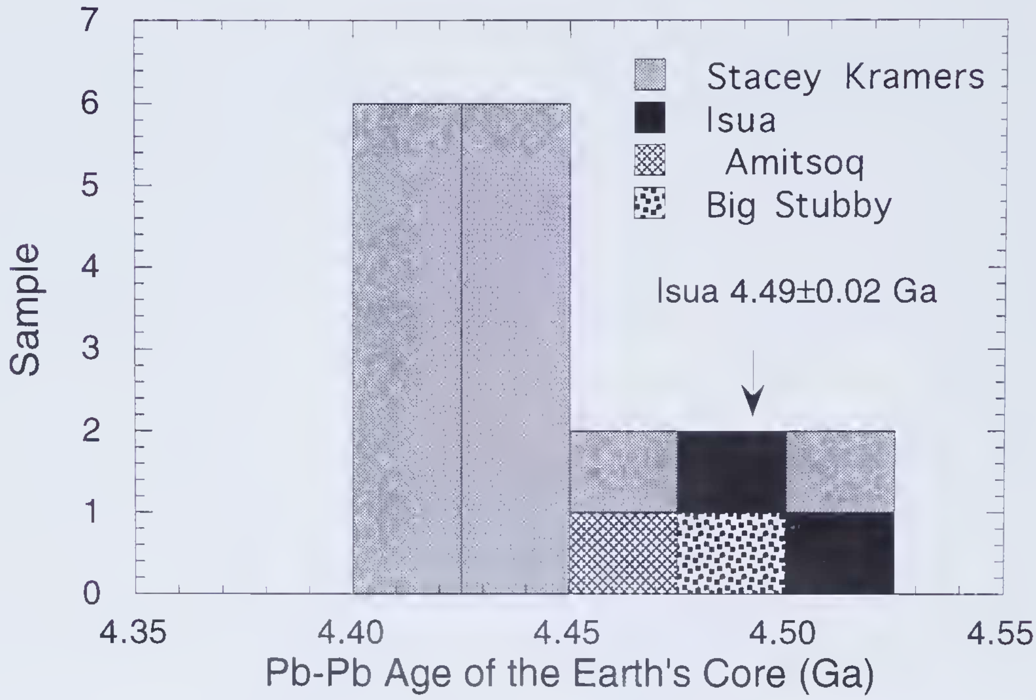

Figure 2. Histogram of single-stage Pb-Pb ages for the Earth's core calculated using data points shown in Fig 1. Arrow shows the best estimate for the mean age of formation of the Earth's core based on Pb-Pb ages from Isua and Big Stubby.

and an equivalent equation for $^{207}Pb/^{204}Pb(1)$. With this minor correction to the Earth's initial Pb and with $\mu_1 = 1$, (the μ from 4.56 Ga to 4.50 Ga) the best estimate of the age of the Earth's core using equation 1, is $T_{core} = 4.49 \pm 0.02$ Ga. This latter estimate is based on both the Isua as well as Big Stubby initial Pb's and is consistent with $\mu_2 = 9$ (the μ from 4.49 Ga to the present). It is noted that the initial Pb from the Amitsoq feldspar is not compatible with this data set, which may be due to the effects of metamorphic disturbance as discussed by Gancarz & Wasserburg (1977) and/or uncertainties in the crystallisation age of the Amitsoq feldspar sample.

Using the same approach, T_{core} ages have also been calculated for the set of younger conformable galena Pb's listed by Stacey & Kramers (1974). A histogram of T_{core} ages is shown in Figure 2 with ages of from 4.40 Ga to 4.50 Ga; these ages are somewhat younger than those from Isua and Big Stubby. Taken at face value, they are indicative of a longer interval (100-200 Ma) for core formation, however the younger galena have significantly more evolved Pb compositions and therefore their interpretation is more problematic.

Sr isotopic constraints on timescales for the accretion of the Earth

Based on the same reasoning as outlined for Pb, the most rigorous constraints on the Earth's initial Sr are those derived from the most primitive terrestrial Sr isotopic compositions. With this in mind, McCulloch (1994) reported a *measured* $^{87}Sr/^{86}Sr$ initial ratio of 0.700502 ± 9 for the 3.46 Ga barite from the North Pole dome of Western Australia. This result, together with a less precise initial Sr ratio for a basaltic komatiite from the Onverwacht Group (Jahn & Shih 1974), indicates an initial Sr isotopic composition for the early Archean (3.46 Ga) mantle of 0.70050 ± 2. This determination currently represents the most reliable value for the early Archean mantle and has important ramifications for the origin of the Earth-Moon system as well as the evolution of the Archean mantle. The key question is how accurate and how representative is this value for the early Archean mantle? Barite is a useful recorder of $^{87}Sr/^{86}Sr$ ratios, but due to its hydrothermal origin may provide only an upper limit to the Sr composition of the Earth's mantle, although in the early Archean the difference in Sr isotopic composition between seawater and the upper mantle is probably insignificant (McCulloch 1994). The consistency of the Pilbara barite result with the few available reliable determinations of initial Sr in the Archean mantle (Jahn & Shih 1974; Machado *et al.* 1986) together with the presence of primitive Pb in the nearby Big Stubby galena (Pidgeon 1978; Richards 1986) indicates that this is a plausible result.

The Earth's initial Sr ratio (BEBI = Bulk Earth Best Initial) can be calculated from the early Archean barite using the following single-stage model:

$$BEBI = (^{87}Sr/^{86}Sr)_{barite} - (^{87}Rb/^{86}Sr)_{EARTH} (e^{\lambda T_{acc}} - e^{\lambda t}) \qquad (3)$$

where $T_{acc} = 4.50$ Ga is the mean age for of accretion of the Earth, t = 3.46 Ga, the age of the Pilbara barite (McNaughton *et al.* 1993), $\lambda = 0.0142Æ^{-1}$ and $(^{87}Rb/^{86}Sr)_{EARTH}$ the ratio from T_{acc} to t. Using a value for $(^{87}Rb/^{86}Sr)_{EARTH}$ of 0.07 ± 0.007 (McCulloch 1994), which reflects the partially depleted character of the early Archean mantle, equation 3 gives a well defined value of BEBI = 0.69940 ± 10. This estimate is significantly greater than that previously assumed using the initial Sr composition of achondrites (*e.g.* Angra dos Reis (ADOR) with $^{87}Sr/^{86}Sr = 0.69893 \pm 2$). Moreover, BEBI is significantly greater than the initial Sr ratio for the Moon (Nyquist *et al.*

1973; Alibert *et al.* 1994) of LUNI = 0.69900 ± 2 and thus allows chronological constraints to be placed on the first 100 Ma of evolution of the Earth-Moon system.

If we consider ADOR as a zero reference time-line for the accretion of planetary bodies, then the difference in the initial Sr compositions between ADOR & BEBI can be translated into a formation interval for planetary accretion relative to ADOR using;

$$I_{Sr}(BEBI) - I_{Sr}(ADOR) = \lambda\ \Delta T\ (^{87}Rb/^{86}Sr)_p \qquad (4)$$

where ΔT is the formation interval and $(^{87}Rb/^{86}Sr)_p$ the parent/daughter ratio in the planetary precursor bodies. By far the most significant uncertainty in applying the ΔT-I_{Sr} methodology to planetary accretion is in estimating the $(^{87}Rb/^{86}Sr)_p$ ratio of the terrestrial planetisimals. The use of a chondritic Rb/Sr ratio cannot be justified as the volatile/refractory ratio of the asteroid belt, as sampled by meteorites, is probably not representative of the solar nebular from which the terrestrial planetesimals accumulated (Taylor & Norman 1990). The ratio of volatile/involatile elements would be expected to increase radially outwards from the Sun following the T-Tauri stage, implying a lower $(^{87}Rb/^{86}Sr)_p$ for proto-Earth materials than for chondrites. Unfortunately, the volatile/involatile ratio of the terrestrial planetesimals cannot be directly ascertained as these bodies where presumably swept up and accreted into the Earth-Moon system.

The other factor that needs to be considered is the extent of volatile loss during planetary formation, particularly if this occurs via accretion of only a relatively small number of more massive bodies with large gravitational potential's and consequently high accretional energies. Rb is volatile under reducing conditions at temperatures <1000 0C whereas it is likely that the Earth's outer surface reached temperatures of >1500 °C (Stevenson 1987). With reasonably efficient mixing of the upper portion of the Earth, concomitant with impact-accretion, it is highly likely that a significant fraction of the proto-Earth's volatile budget including Rb (and Pb) would not have been accreted. The actual magnitude of the Rb/Sr (*i.e.* volatile/refractory) fractionation that occurred during accretion of the Earth although obviously significant, is difficult to quantify. An upper limit to the extent of fractionation is given by the difference between the bulk Earth and chondritic $^{87}Rb/^{86}Sr$ ratios of 0.082 and ≈0.8 respectively. This indicates a maximum possible fractionation of Rb/Sr resulting from a combination of both planetary accretion and Solar nebular processes of ≈10x. An intermediate estimate is given by $(^{87}Rb/^{86}Sr)_p$ = 0.4 which implies a depletion of Rb/Sr in the terrestrial region of the solar nebular of ≈x2 compared to chondrites. For BEBI = 0.69940, this corresponds to ΔT of ≈80 Ma relative to ADOR (Fig 3). Substantially greater depletion in the volatile content of the terrestrial plan-

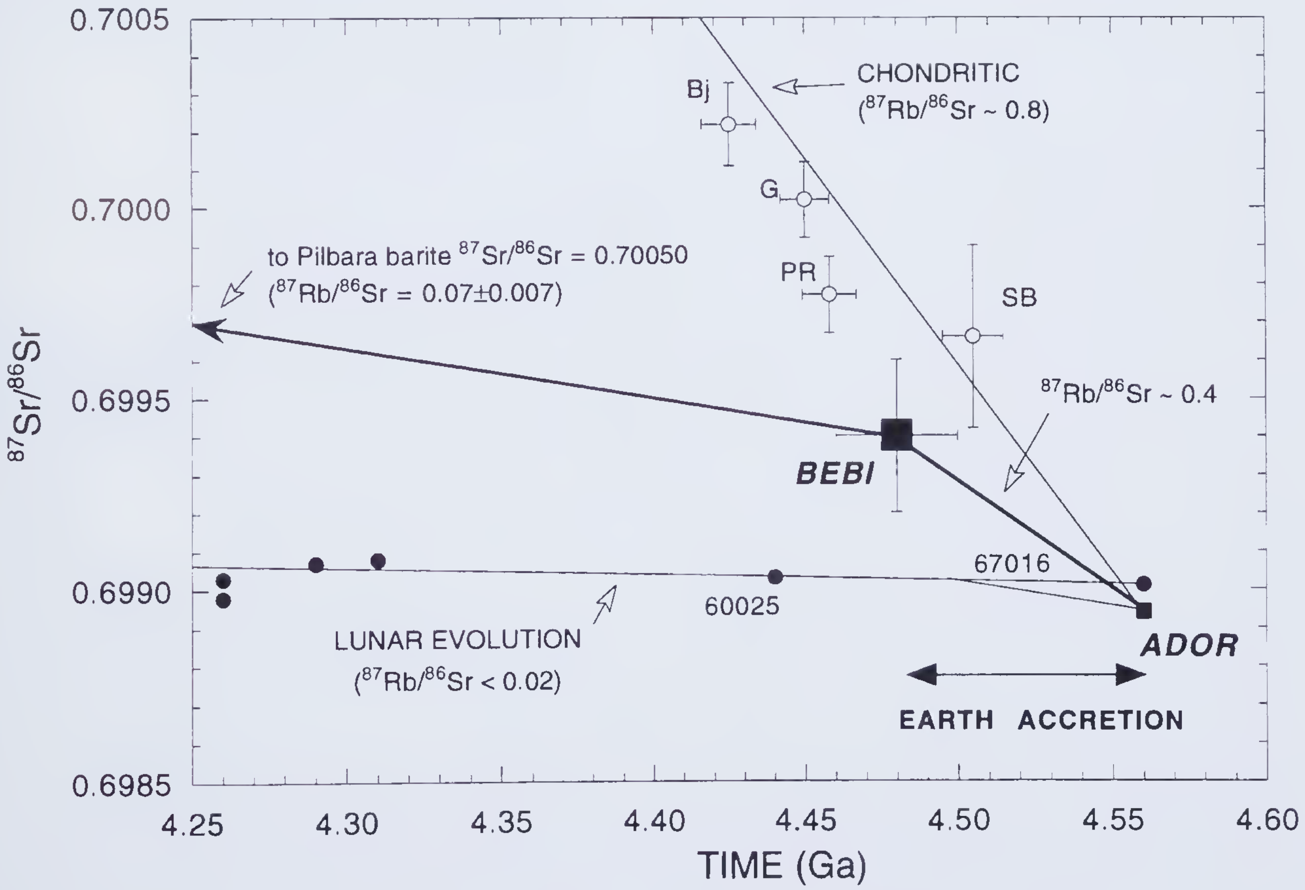

Figure 3. Plot of initial Sr composition versus time for the Earth = BEBI (McCulloch 1994), Moon shown with solid circles (Alibert *et al.* 1994) and ordinary chondrites with open symbols (Brannon *et al.* 1987), relative to the primitive differentiated achondrite Angra dos Reis = ADOR (Lugmair & Galer 1992; McCulloch 1994). Bj = Bjurbole, G= Guarena, PR = Peace River, SB = Soko Banja, (Brannon *et al.* 1987). The mean age for accretion of the Earth ranges from ≈80 Ma to 100 Ma for a $^{87}Rb/^{86}Sr$ ratio of ≈0.4, which assumes an ≈50% loss of volatiles during accretion.

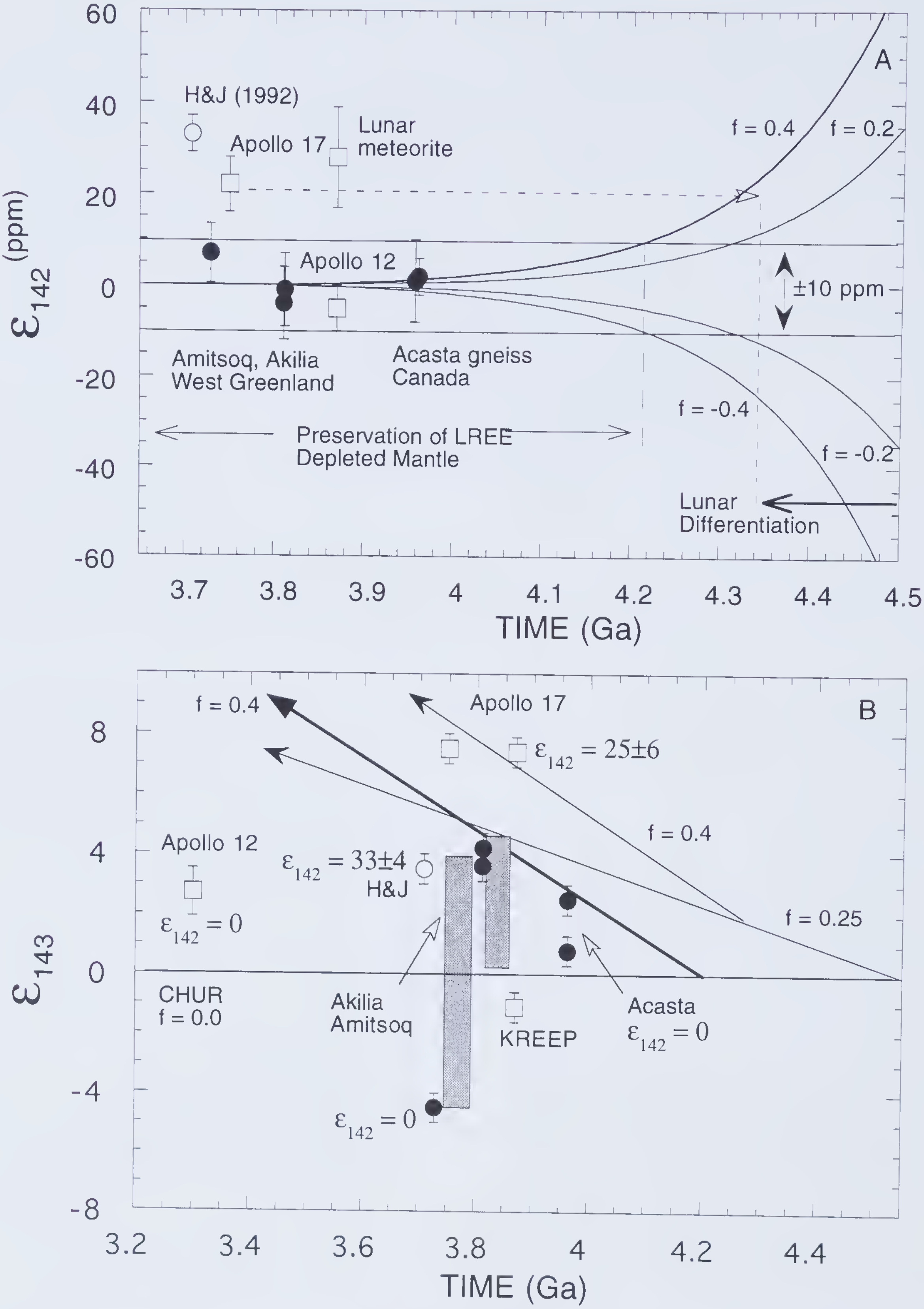

Figure 4A. Plot of ε_{142} versus time for an initial $^{146}Sm/^{144}Sm$ = 0.007 (Prinzhofer *et al.* 1992). If a LREE depleted mantle (f > 0.2) formed in the first 200-300 Ma of Earth history then within the analytical resolution of ± 10 ppm it would be expected to preserve positive ε_{142} values. The general absence of terrestrial ε_{142} excesses (McCulloch & Bennett 1993) therefore suggests that fractionated terrestrial reservoirs were not preserved until after ≈4.3 Ga, after which, ^{146}Sm was extinct. Only a single positive ε_{142} value of 33 ± 4 ppm, has so far been reported by Harper & Jacobsen (1992). Lunar data is from Nyquist *et al.* (1995). **B.** Plot of ε_{143} versus time for earliest Archean rocks from Greenland (Amitsoq and Akilia are shaded regions) and Acasta gneiss from Canada (Bennett *et al.* 1993). Despite the large range of both positive and negative terrestrial ε_{143} values, there is general absence of associated terrestrial ε_{142} effects (samples shown in solid symbols) apart from that reported by Harper & Jacobsen (1993). This indicates that the terrestrial record of highly fractionated mantle reservoirs was not generally preserved until after ≈4.3 Ga. The Apollo 17 samples (Nyquist *et al.* 1995) with very positive ε_{143} values show relatively small ε_{142} effects of 25 ± 6 ppm (Fig 4A), consistent with crystallisation of the lunar magma ocean at ≈4.3 Ga.

etesimals is probably unlikely, as this would give $\Delta T >$ 100 Ma, which would correspond to an age for the Earth that is younger than that obtained for terrestrial core formation ($\approx$4.49 Ga) using Pb-Pb isotope systematics outlined previously.

Sm-Nd constraints on the Earth's age and early differentiation

In contrast to the U-Pb and Rb-Sr systems, the rare earth elements Sm and Nd are refractory elements that are strongly fractionated only during magmatic differentiation processes. The Sm-Nd isotopic system is however unique in having both extinct and "live" parent-daughter decay systems; the now extinct ^{146}Sm-^{142}Nd with a half-life of 1.03 10^{8} yrs, and the commonly used long-lived ^{147}Sm-^{143}Nd system with a half-life of 1.06 10^{11} yrs. Owing to its short half-life, ^{146}Sm effectively became extinct at $\approx$4.30 Ga, at which time the ^{142}Nd/^{144}Nd ratio became essentially invariant. Positive or negative deviations in the observed ^{142}Nd/^{144}Nd ratio must therefore represent the effects of differentiation processes operative prior to 4.30 Ga. Thus ^{142}Nd isotopic abundances have the potential to reveal events occurring in the first 250 million years of the Earth history.

The combined ^{142}Nd-^{143}Nd isotopic systematics can potentially provide constraints on the magnitude and the longevity of chemical differentiation in the early Earth due, for example, to the presence of an early magma ocean. Positive or negative deviations of ε_{142} values from the bulk Earth value are possible if a REE fractionated reservoir was formed within the first 100 to 200 Ma of Earth history and if this reservoir survived intact as a closed system until it could be sampled via magmatic processes and preserved in the oldest terrestrial rocks (3.70 - 3.96 Ga). This scenario is shown in Figure 4A where ε_{142} values are plotted versus the time of reservoir fractionation and isolation. In these calculations, it is assumed that the Solar System had ^{146}Sm/^{144}Sm = 0.007 at 4.56 Ga, the same Sm isotopic composition as that determined from meteorites (Prinzhofer *et al.* 1992; Nyquist *et al.* 1995). The magnitude of any ε_{142} anomaly is directly dependent on the fractionation (f) where $f = [(Sm/Nd)/(Sm/Nd)_{CHUR} - 1]$, as well as the time T when the reservoir fractionated and became isolated. Following Harper & Jacobsen (1992), the ε_{142} effects can be calculated from the following relationship;

$$\varepsilon_{142}(t) = f\, Q_{142}[^{146}Sm/^{144}Sm][e^{-\lambda_{146}(T_0-T)} - e^{-\lambda_{146}(T_0-t)}] \qquad (5)$$

where Q_{142} = 354 Ga^{-1}, T_0= 4.56 Ga, T = time of mantle differentiation, and t = crystallisation age of the rock.

Apollo 17 lunar basalts (Nyquist *et al.* 1995) show small ε_{142} effects of $\approx$25 ppm (Fig 4) together with very positive ε_{143} values of >+7 at 3.9 Ga (Fig 4B). This is consistent with crystallisation of the lunar magma ocean at $\approx$4.3 Ga with f values of from 0.25 to 0.4 (Fig 4B). The terrestrial samples analysed by McCulloch & Bennett (1993) are consistent with the lunar results of Nyquist *et al.* (1995), having ε_{143} values $\approx$1/2 that of the Moon and no measurable ε_{142} effects >10 ppm. Harper & Jacobsen (1992) have however reported an ε_{142} of 33 ppm value for a metasedimentary sample from Isua, with ε_{143} = 3.5, the latter being significantly smaller value than that found in Apollo 17 basalts. Thus at this time, the Harper & Jacobsen (1992) result must be considered anomalous with respect to both the Earth and Moon and not representative of the terrestrial early Archean mantle.

The strongly positive ε_{143} values shown in Figure 4B for terrestrial samples together with the *lack* of ε_{142} effects (McCulloch & Bennett 1993) provides a firm upper limit to the time of differentiation of the terrestrial mantl. These constraints require that highly fractionated LREE depleted reservoirs were formed and only preserved sometime after 4.3 Ga, by which time decay of ^{146}Sm was essentially complete. In the interval from $\approx$4.5 Ga, to 4.3 Ga the Earth's mantle was probably well mixed and constantly being rehomogenised, possibly due to the effects of giant impacts. The Nd isotopic results are therefore consistent with a relatively extended interval ($\approx$100 Ma) for the formation of the Earth.

Discussion and Conclusions

The interpretation of ΔT's determined using the Pb-Pb and Rb-Sr chronologies is not straightforward. Essentially, the Pb-Pb system records the integrated history of core formation while Rb-Sr records the volatile/refractory fractionation with neither system discriminating between various rates of core formation or accretion. Thus the ΔT inferred for the Earth of $\approx$100 Ma does not necessarily imply that the Earth accreted continuously throughout this interval. For example, the Earth may have formed essentially catastrophically in a period of <10^7 yrs, but $\approx$80 to 100 Ma following the differentiation of ADOR. Alternatively, if accretion and core formation had been essentially continuous throughout this period, then the formation interval would have been significantly longer than the mean ages given by the Pb-Pb and Rb-Sr isotopic systems. In the latter case, the mean age for accretion and core formation of $\approx$4.49 Ga, implies a formation interval of $\approx$140 Ma (Fig 5). Thus a formation interval for the Earth of $\approx$100 Ma is a relatively conservative lower estimate.

What are the implications of a relatively prolonged formation interval for the Earth on the origin of the Moon? Assuming that the Earth and Moon were derived from similar precursor materials, then to account for the Moon's lower initial Sr ratio (Alibert *et al.* 1994), it is required to have formed first, probably via a giant impact on the proto-Earth, within $\approx$20 Ma of ADOR. The strong terrestrial geochemical signature of the Moon (Ringwood 1986) would have been inherited from the proto-Earth at this time, indicating that the Earth's core had started to form. Immediately following the giant impact, both the proto-Earth and proto-Moon are molten, but if large-scale differentiation occurs on the Earth, its effects were shortlived due to ongoing accretion and thorough mixing. From $\approx$4.54 Ga to $\approx$4.48 Ga, the Earth was rehomogenised while continuing to accrete the remaining $\approx$30% of its final mass. Core formation probably occurred concomitantly with accretion rather than being triggered by a late-stage giant impact. Both the terrestrial Sr, Nd and Pb-Pb isotopic systematics are thus consistent with an extended ($\approx$100 Ma) interval for the accretion of the Earth.

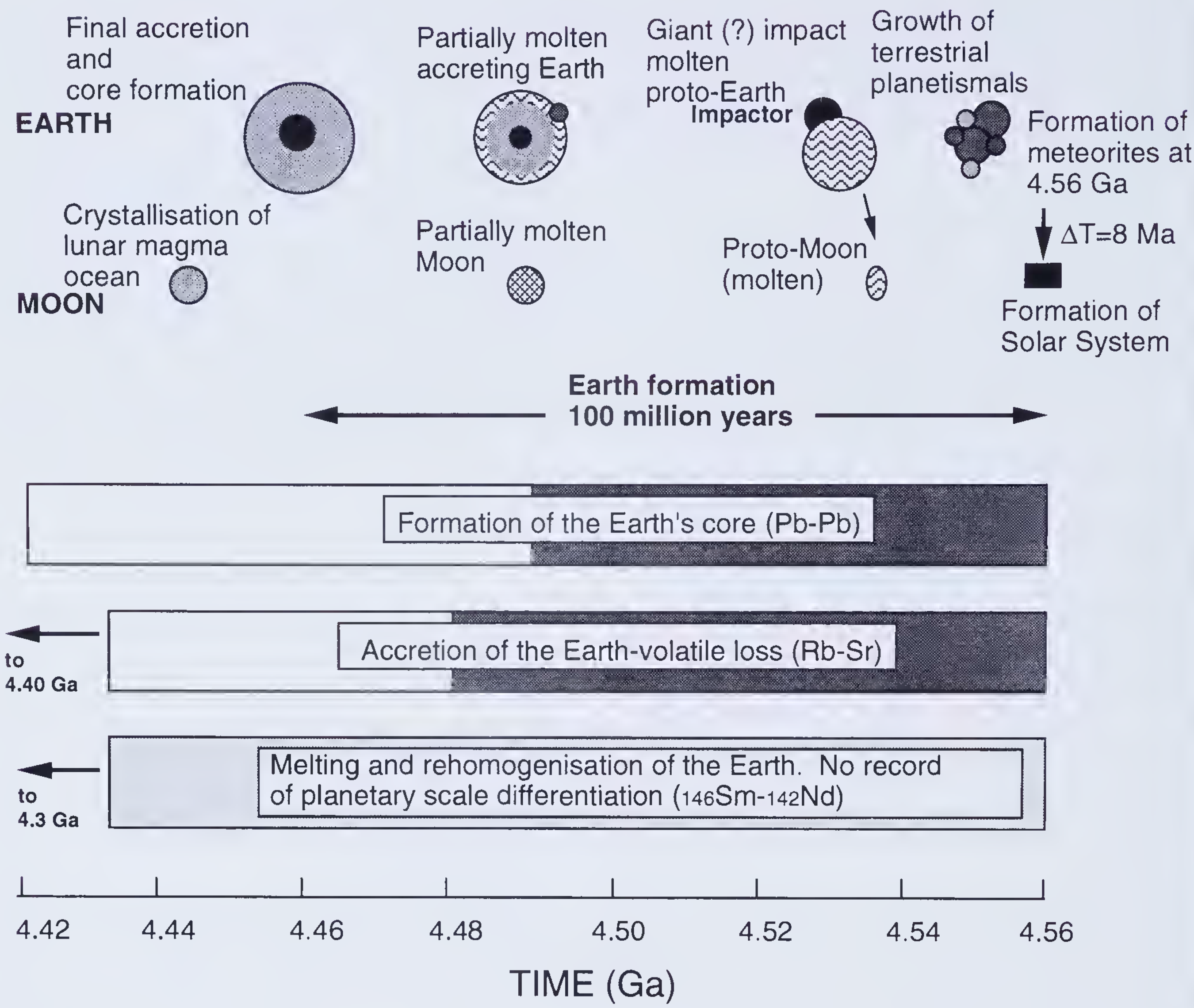

Figure 5. Schematic diagram illustrating the sequence of events during accretion and early differentiation of the Earth-Moon system. Meteorites first formed at 4.563 Ga with an ≈8 Ma range of ages (Allegre *et al.* 1995). The mean age for core formation and accretion of the Earth is 4.49 ± 0.02 Ga and 4.48 ± 0.04 Ga respectively (solid shading). Accretion and core formation may have continued until ≈4.42 Ga (open boxes) consistent with the Earth forming over at least an ≈100 Ma interval. Constraints from ε_{142}-ε_{143} values indicate that the differentiation of the mantle into distinctive long-lived reservoirs did not commence until after ≈4.3 Ga, consistent with rehomogenisation of the Earth's mantle during the first ≈200-300 Ma of its history.

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